ATMOSPHERIC WATER VAPOR TRANSPORT AND THE WATER BALANCE OF NORTH AMERICA

II. LARGE-SCALE WATER BALANCE INVESTIGATIONS

EUGENE M. RASMUSSON

Geophysical Fluid Dynamics Laboratory, ESSA, Washington, D.C.

ABSTRACT

The atmospheric water vapor flux divergence and certain aspects of the water balance of North America are investigated, using data from the period May 1, 1958–Apr. 30, 1963.

The mean vertical distribution of flux divergence is computed for the United States for the months of January (1962, 1963) and July (1961, 1962). Strong flux convergence in the lowest kilometer and divergence in the remainder of the troposphere were found in July. Flux convergence was found throughout the troposphere over the eastern half of the area in January, with a maximum between 900 and 950 mb.; while in the west, convergence (with no particularly pronounced maximum) was found above 800 mb., with weak divergence below. Corresponding features of the profiles were found at higher elevations over the west, where the flux divergence above 500 mb. is quite significant.

Particular emphasis is placed on computation of the vertically integrated flux divergence, and its use in estimating $\overline{P-E}$, the mean difference between precipitation and evapotranspiration. As in the case of the flux field, the flux divergence exhibits a pronounced diurnal variation south of 50°N., particularly during the summer. Nevertheless, the results of water balance computation over the United States and southern Canada, using twice-daily observations from the existing aerological network, indicate that reliable mean annual, season, and monthly values of $\overline{P-E}$ can usually be obtained when averaging over areas of 20×10^5km.^{2} or larger. Averages over smaller areas are less reliable, and become quite erratic as the size of the area is reduced to less than 10×10^5km.^{2} This deterioration is mainly due to the presence of a systematic error pattern of relatively large scale and amplitude.

The mean monthly values of evapotranspiration and storage change, obtained from balance computations over the United States and southern Canada, and over that portion of the area east of the Continental Divide are presented and discussed. A comparison of values of evapotranspiration computed by means of the atmospheric water vapor balance equation, with those computed using Thornthwaite climatic water balance data indicates that over the United States and southern Canada the latter systematically overestimates $\overline{P-E}$ during the winter, and underestimates it during the summer by a substantial amount. This contributes to a computed seasonal change in surface and subsurface storage which averages more than twice that obtained from an evaluation of the flux divergence.

Examination of the relationship between precipitation and storage over eastern North America indicates that for areas of this size, the departure from normal of precipitation by itself serves as a fairly good quantitative indicator of the departure from normal storage change.

CONTENTS

1. INTRODUCTION

An accurate quantitative knowledge of the components of the hydrologic cycle of the earth-atmosphere system, on a regional and global basis, is of basic importance in many branches of geophysics. Progress toward such knowledge has, however, been seriously hindered by inadequate measurement of many of the processes involved (Ackerman [1]). Information is still inadequate on soil moisture storage, ground water storage and movement, and in many areas, precipitation. In addition, estimates of average evapotranspiration over large areas is still far from satisfactory. Ackerman [1] states: "Changes in regional or global supply of atmospheric moisture obtained from land and water surfaces by evapotranspiration processes are largely unknown . . . new instruments or improved techniques for use with conventional instrumentation are needed . . . to quantify the exchange of moisture with the atmosphere over large areas for which water balance evaluations are required."

Even a simple balance equation for the terrestrial branch of the hydrologic cycle normally contains two unmeasured quantities, evapotranspiration and change in surface and subsurface storage. Thus one must rely on some additional relationship in order to solve for the unknowns. The conventional approach to this problem has centered on attempts to estimate actual evapotranspiration through the use of some additional empirical relationship which relates evapotranspiration to measured surface meteorological parameters, and to the soil moisture deficit. Soil moisture storage changes are then computed as a residual from the terrestrial balance equation. Such techniques have been discussed by Thornthwaite and Hare [27] and Kohler and Richards [12]. Changes in lake and stream storage can be estimated, but the quantitative changes in ground-water storage over large areas is, for the most part, unknown.

There is a further problem involved in the use of the terrestrial water balance equation which is sometimes overlooked. This arises from the fact that errors in the measurement of precipitation are not random, but exhibit a negative bias (LaRue and Younkin [13]). Consequently, measured precipitation will, in most cases, be less than actual precipitation. This bias, which has been the subject of numerous investigations during the past century, is thoroughly discussed in the comprehensive survey of Weiss and Wilson [29]. The error is mainly related to the speed of the wind and the character of the precipitation, and is accentuated in mountainous areas where reports are often sparse and biased toward lower elevations.

Added problems arise in the measurement of snow. Struzer, Nechayev, and Bogdanova [24], in a study of precipitation measurements over the U.S.S.R., found errors in the measurement of mean annual precipitation which varied between approximately 5 and 50 percent, depending on the climatic zone. Errors in the measurement of mean monthly precipitation during the winter months were sometimes in excess of 100 percent.

The average amount by which precipitation is underestimated over North America is, of course, difficult to say. Even a figure as low as 5 to 10 percent, which would not seem unreasonably high, amounts to an average of 3.5 to 7.5 cm./yr. over the United States. This is by no means a negligible figure when considering long-term storage changes and, as the results of this study will show, probably represents at least as large a systematic error as that which arises in the evaluation of the mean vapor flux divergence over the United States and southern Canada east of the Continental Divide. Since precipitation measurements were used in this investigation in order to obtain estimates of evapotranspiration, the computed values of evapotranspiration will show the same negative bias as the measured precipitation.

A number of studies during the past several years, including those of Benton and Estoque [3], Hutchings [10], Starr and Peixoto [22], Palmén [17], Lufkin [14], and Starr, Peixoto, and Crisi [23] have demonstrated that the atmospheric vapor flux divergence can in many cases be measured accurately enough to give useful estimates of the mean difference between evapotranspiration and precipitation. In order to obtain satisfactory results, an adequate aerological network must exist, the region considered must not be too small, and the time period over which the observations are averaged must be of sufficient length to render the effect of random errors negligible. Under such conditions one can use the atmospheric vapor balance equation and the terrestrial balance equation as independent relationships from which to evaluate evapotranspiration and total storage change, that is, the conbined change in surface and subsurface storage. There is, in theory, a particular advantage in using this technique for the evaluation of storage changes, since the use of measured values of precipitation and empirically computed values of evapotranspiration is entirely avoided.

The large-scale characteristics of the vapor flux field over North America and the Central American Sea were described in Part I of this paper (Rasmusson [19]). The balance equations used for the budget computations and a description of the data and procedures used in these computations will be presented in Part II. This will be followed by a discussion of some significant characteristics of the vertical distribution of the vapor flux divergence. Systematic errors which were encountered in the divergence computations will be described. It was noted in Part I that the vapor flux exhibited significant diurnal variations. This is also true in the case of the flux divergence, and the characteristics of these variations will be illustrated and discussed.

Finally, the results of balance computations for some of the larger areas investigated will be discussed. Computations for the United States and southern Canada cover a 2-yr. period (May 1, 1961–Apr. 30, 1963). For that portion of the United States and southern Canada east of the Continental Divide, which was chosen for the most detailed study, the computations have been extended to 5 yr. (May 1, 1958–Apr. 30, 1963). The extension of the period of investigation from 2 to 5 yr., completed since the publication of Part I of this paper, has yielded results which are considerably more definitive than those previously obtained, particularly for the smaller subdivisions of the area.

2. THE BALANCE EQUATIONS

The following notation will be used:

g = acceleration of gravity

a = mean radius of the earth

λ = longitude

ϕ = latitude

p = pressure

q = specific humidity

p_s = pressure at the ground

p_u = pressure above which the vapor flux divergence becomes negligibly small

$u = a \cos\phi \dfrac{d\phi}{dt}$, zonal wind component

$v = a \dfrac{d\phi}{dt}$, meridional wind component

$\mathbf{i}_\lambda, \mathbf{i}_\phi$ = eastward- and northward-pointing unit vectors, respectively

$\mathbf{G}$ = total subsurface flow (gm. (cm. sec.)$^{-1}$)

E = rate of evapotranspiration

P = rate of precipitation

R_o = rate of stream flow from a drainage area

S = total water storage on and below the surface of the earth per unit horizontal area

Σ = net sources of water vapor in a unit atmospheric column extending from p_s to p_u

t, τ = time

C = curve bounding a drainage area

$\mathbf{n}_c$ = outward-pointing unit normal vector on curve

$\overline{(\ )} = \dfrac{1}{\tau}\displaystyle\int_\tau (\ )dt$ = time mean

$(\ )' = (\ ) - \overline{(\ )}$ = instantaneous departure from time mean

$$\langle(\ )\rangle=\frac{1}{A}\iint_A(\ )a^2\cos\phi\, d\lambda\, d\phi=\text{spatial mean.}$$

The following vertical integrals will be referred to in the course of this discussion:[1]

$\overline{W}=\frac{1}{g}\int_{p_u}^{p_s}\overline{q}\,dp$ mean precipitable water (gm. cm.$^{-2}$ or cm.)

$\overline{Q}_\lambda=\frac{1}{g}\int_{p_u}^{p_s}\overline{qu}\,dp$ vertically integrated mean total zonal water vapor flux (gm. [cm. sec.]$^{-1}$)

$\overline{Q}_\phi=\frac{1}{g}\int_{p_u}^{p_s}\overline{qv}\,dp$ vertically integrated mean total meridional water vapor flux (gm. [cm. sec.]$^{-1}$)

$\overline{\mathbf{Q}}=\mathbf{i}_\lambda Q_\lambda+\mathbf{i}_\phi Q_\phi$ vertically integrated mean total water vapor flux (gm. [cm. sec.]$^{-1}$).

The form of the atmospheric water vapor balance equation is essentially that of Starr and Peixoto [22].

Assuming hydrostatic equilibrium, one may write the atmospheric water vapor balance equation for a column of air, extending from the ground to a pressure p_u, in the following form:

$$\frac{\partial W}{\partial t}+\nabla\cdot\mathbf{Q}=\Sigma. \tag{1}$$

Evapotranspiration from the earth's surface and precipitation falling from the air column constitute the major source and sink of water vapor. The formation (evaporation) of clouds within the column constitutes another possible sink (source), but the use of commonly accepted values for the water content of clouds (aufm Kampe and Weickmann [11]; Atlas [2]) indicates that the flux of water, in liquid and solid form, will rarely average 10 to 20 gm. (cm. sec.)$^{-1}$ for periods of a month or more. This, for example, represents around 1 percent of the mean flux in the regions of persistent wintertime cloudiness along the west coast of North America. Since the flux divergence rather than the flux itself affects the accuracy of the water balance computation, it can be concluded that the transport of water in liquid or solid form may be of significance in those relatively localized regions of persistent formation or dissipation of clouds, or for occasional short time periods, but can normally be ignored on a mean monthly basis for large-scale water balance studies.

Thus

$$\Sigma=E-P.$$

When applied to mean conditions over a given region and time period, equation (1) becomes

$$\langle\nabla\cdot\overline{\mathbf{Q}}\rangle=\overline{\langle E-P\rangle}+\left\langle\frac{\partial\overline{W}}{\partial t}\right\rangle. \tag{2}$$

For annual means, $\langle\overline{\partial W/\partial t}\rangle$ is usually negligible compared with the other terms. For monthly means, however, all terms are often of the same order of magnitude, particularly during the spring and fall.

[1] The author would like to take this opportunity to correct an error on page 404 of Part I of this paper [19]. Under "Data and Procedures," the second line of equations should read:

$$\overline{qv}=\frac{\Sigma qv}{N},\ \overline{q'u'}=\overline{qu}-\overline{q}\,\overline{u},\ \overline{q'v'}=\overline{qv}-\overline{q}\,\overline{v}.$$

Also, on page 424, the dates in the third line of the "Summary" should be "May 1, 1961–April 30, 1963."

The vapor flux divergence can be expressed in spherical coordinates:

$$\nabla\cdot\overline{\mathbf{Q}}=\frac{1}{a\cos\phi}\left\{\frac{\partial\overline{Q}_\lambda}{\partial\lambda}+\frac{\partial(\overline{Q}_\phi\cos\phi)}{\partial\phi}\right\}. \tag{3}$$

This expression can be conveniently evaluated by finite difference methods to provide the mean divergence within each area defined by 4 grid points. However, when making detailed water balance studies which involve the use of streamflow data, it is usually more convenient to obtain the mean divergence over an irregularly shaped drainage basin. For this purpose, application of the Gauss Theorem gives:

$$\langle\nabla\cdot\overline{\mathbf{Q}}\rangle=\frac{1}{A}\oint_c\overline{\mathbf{Q}}\cdot\mathbf{n}\,dc. \tag{4}$$

A second relationship is obtained as a balance equation for the ground branch of the hydrologic cycle. When applied to a particular drainage basin, this balance may be expressed, in its simplest form, as follows:

$$-\overline{\langle E-P\rangle}=\langle\overline{R}_o\rangle+\left\langle\frac{\partial\overline{S}}{\partial t}\right\rangle\frac{1}{A}\oint_c\overline{\mathbf{G}}\cdot\mathbf{n}_c dc. \tag{5}$$

$\langle\overline{R}_o\rangle$ is the net stream outflow from the basin, and $\langle\overline{\partial S/\partial t}\rangle$ is the mean rate of storage change (surface, soil moisture, and ground water) over the basin. $\oint_c\overline{\mathbf{G}}\cdot\mathbf{n}_c dc$ is the net underground flow through the vertical boundaries of the basin. Note that ground water which discharges into streams within the basin does not contribute to this term; the term is nonzero only when ground water and surface divides do not coincide. Most underground exchange between North American drainage basins probably occurs on a scale too small to be studied to advantage using the atmospheric water vapor balance equation. Lacking evidence to the contrary, such exchanges are assumed to be small over the large drainage areas investigated, when compared with the seasonal and interannual surface and subsurface storage changes.

Neglect of the underground exchange term then leaves only two unknowns, $\overline{\langle E-P\rangle}$ and $\langle\overline{\partial S/\partial t}\rangle$, to be evaluated between equations (2) and (5), since $\langle\overline{\partial W/\partial t}\rangle$ can be measured. Solving for surface and subsurface storage change gives:

$$\left\langle\frac{\partial\overline{S}}{\partial t}\right\rangle=\langle\nabla\cdot\overline{\mathbf{Q}}\rangle+\left\langle\frac{\partial\overline{W}}{\partial t}\right\rangle-\langle\overline{R}_o\rangle. \tag{6}$$

Using precipitation measurements, one can also solve for $\langle\overline{E}\rangle$:

$$\langle\overline{E}\rangle=-\langle\nabla\cdot\overline{\mathbf{Q}}\rangle+\langle\overline{P}\rangle+\left\langle\frac{\partial\overline{W}}{\partial t}\right\rangle. \tag{7}$$

These two simple relationships can then be used to evaluate the two unknowns of the terrestrial water balance equation, all other quantities in the equations being measured.

3. DATA AND PROCEDURES

Aerological data used in this investigation and the analyses of the flux fields were discussed in Part I (Rasmusson [19]).

Vapor flux data used in this study generally extended no higher than 300 mb. Thus, any contribution from the upper 300 mb. of the atmosphere to the vertically integrated vapor flux divergence is not included in the integrals. We may make a rough estimate for an upper bound on the error which is introduced by this data deficiency. Consider a hypothetical square with sides 1000 km. in length. The enclosed area is then 10^6km.^{2}, the order of magnitude of the typical regions of our water balance computations. Assume further a steady west-east current across the area above 300 mb. Next assume a mean upward flux of moisture through the 300-mb. level, at a rate sufficient to increase the vapor content of the air passing horizontally through the 150–300-mb. layer by an average of 0.3 gm./kg. This would appear to be a rather generous moisture increase when one considers the low values of saturation specific humidity of this layer, the size of the area involved, and the fact that this is a mean value for a period of 1 mo. One may then compute the mean west wind speed along the 1000-km. eastern boundary which would be required to remove moisture moving upward through the 300-mb. level at the rate of, say, 1.0 gm. (cm.^{2}mo.)$^{-1}$, i.e., to give a mean divergence contribution from the 300–150-mb. layer of 1.0 gm.(cm.^{2}mo.)$^{-1}$. The mean wind speed required to accomplish this would be around 25 m./sec. Thus for large-scale water balance computations, the order of magnitude of the error arising from neglect of the divergence contribution above 300 mb. could conceivably be as high as 1 gm.(cm.^{2}mo.)$^{-1}$, but a value on the order of 0.1 gm.(cm.^{2}mo.)$^{-1}$ would probably be more realistic.

Computations of flux divergence were made by applying finite difference methods to equation (3) (Peixoto [18]), using as data the values of $\overline{Q}_\lambda$ and $\overline{Q}_\phi$ on a 2.5° lat. by 2.5° long. grid south of 57.5°N., and on a 5.0° long. by 2.5° lat. grid north of this latitude. As before, individual computations were made for each month, and separately for the 12 GMT and 00 GMT data. In order to obtain accurate values of mean divergence for the various irregularly shaped regions considered in the water balance studies, the net flux across a convenient curve, closely approximating the actual boundary of the basin, was estimated directly from the flux component maps.

Bock, Frazier, and Welsh [5], using the same data source, and an analysis program which was run on the UNIVAC 1108 computer at GFDL, have objectively computed the mean monthly flux divergence over North America for the 5-yr. period May 1, 1958, through Apr. 30, 1963. Their results, and those obtained from the hand analyses, were in good agreement over the United States and southern Canada east of the Continental Divide during the 2-yr. period of overlap. Their results were therefore used in obtaining a 5-yr. balance for that area.

Streamflow data were obtained from the water supply papers of the U.S. Geological Survey and Water Resources Papers of the Canadian Department of Northern Affairs, for an area of 85.7×10^5km.^{2} covering almost all of the United States and much of southern Canada. Immediate coastal regions were not included, partly because of the time involved in obtaining runoff from the large number of small coastal streams and partly due to limitations imposed by the location of the last downstream stream-gaging station, which is normally some distance inland. This had the effect of keeping the boundary of the drainage area well within the outer ring of aerological stations. A listing of stream-gaging stations used in this study, the areas they gage, and additional regions of internal drainage included in the area can be found in a previous report (Rasmusson [20]).

The accuracy of streamflow data depends primarily on 1) the stability of the stage-discharge relation or, if the stream channel is unstable, the frequency of the discharge measurements; and 2) the accuracy of observations of stage, measurements of discharge, and interpretation of records (U.S. Geological Survey Water Supply Papers). The station description states the degree of accuracy of the records. The error in daily values is generally less than 10 percent; consequently the mean monthly and annual errors will, in general, be considerably less than this figure. The author knows of no systematic errors in these data. For a more complete discussion of stream-gaging procedures, quality of data, and streamflow characteristics, see Roden [21].

Precipitation data were obtained from U.S. Weather Bureau State Climatological Summaries, and the Monthly Report of the Canadian Department of Transport.

With regard to the aerological data, several stations, mostly military operated, converted from the lithium chloride to the carbon humidity element during this period. A study of these data (Rasmusson [20]) indicated no large differences in the monthly mean flux as measured by the different elements. For a discussion of the representativeness of the vapor flux data and analyses, the reader is referred to Rasmusson [20].

4. THE VERTICAL DISTRIBUTION OF FLUX DIVERGENCE

A general investigation of the vapor flux divergence on various pressure surfaces was not attempted in this study. However, the series of vertical cross sections illustrated in Part I of this paper and cross sections at 100°W. were used to define the flux through the boundaries of two areas. The first was bounded on the south and north by 30°N. and 47.5°N., and on the east and west by 80°W. and 100°W. The second area extended from 80°W. to the Pacific Coast and was bounded on the south by 30°N. east of 105°W. and 32.5°N. west of 105°W., and on the north by 47.5°N. Except for that portion of the boundary between 30° and 32.5°N. at 105°W., the flux was completely depicted on the cross sections.

Values of the boundary flux were tabulated at 50-mb. intervals from 1000 mb. to 400 mb. from data on the cross sections. Data for El Paso were used as an estimate of the zonal flux through the gap at 105°W. Additional values were interpolated from the cross sections at 975 mb. and 925 mb. when needed to properly define the vertical profiles. The ground profile along the boundaries was estimated as accurately as possible and transport was computed only where the pressure surface was above ground level.

The total outflow at each pressure surface was divided by the enclosed area (eastern region 33.6×10^5 km.^{2};

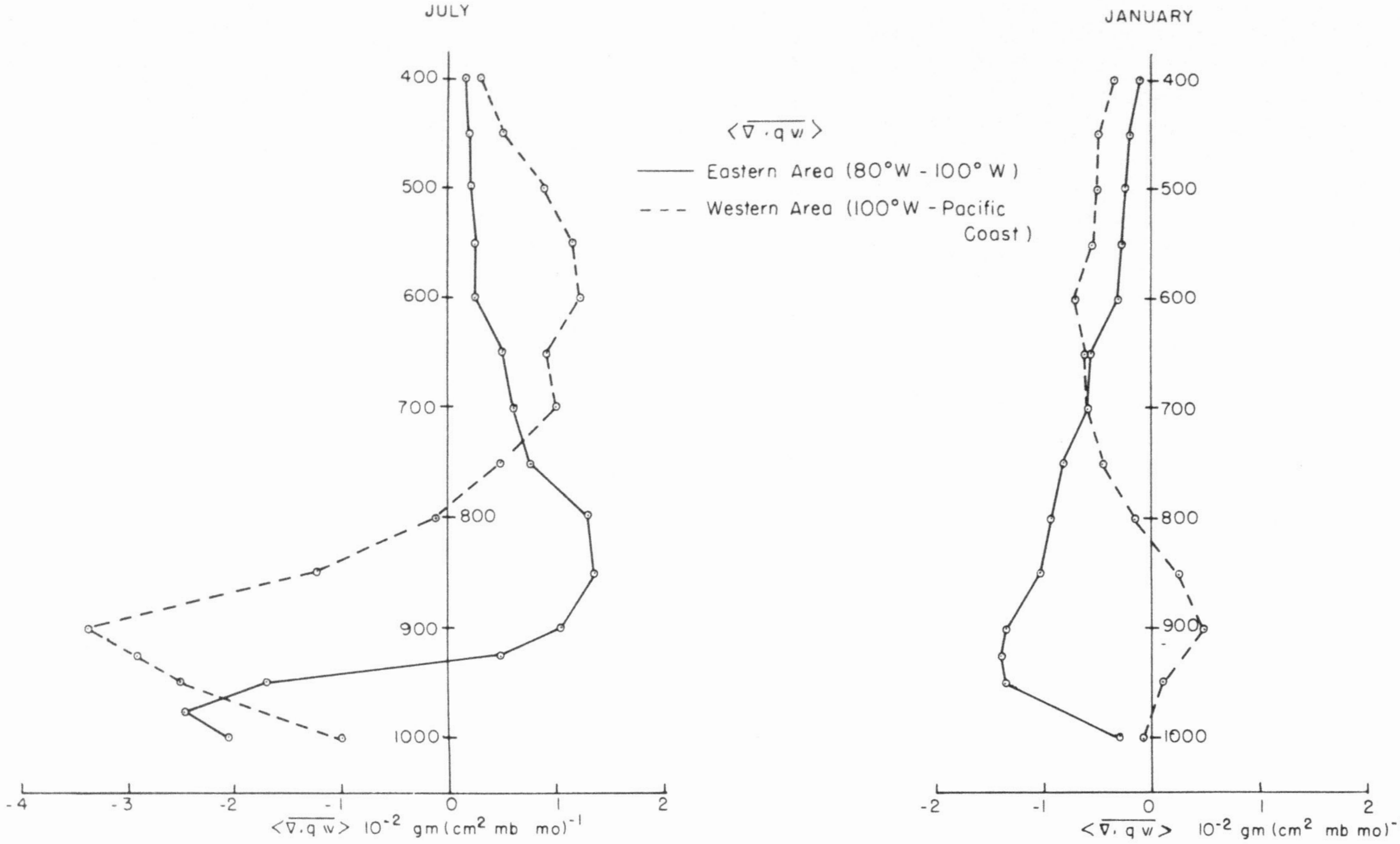

FIGURE 1.—Vertical distribution of $\langle \nabla \cdot \overline{q\mathbf{v}} \rangle$, January 1962, 1963, and July 1961, 1962. The western area is bounded on the south by 30°N. (100°W.–105°W.) and 32.5°N. (105°W.–Pacific coast); on the north by 47.5°N. The eastern area is bounded by lat. 30°N. and 47.5°N. and long. 80°W. and 100°W. $\langle \nabla \cdot \overline{q\mathbf{v}} \rangle$ is an average of 00 GMT and 12 GMT observations over the total enclosed area.

western region 32.6×10^5 km.^{2}) in order to obtain a value of outflow per unit horizontal area. This value is equivalent to the areal mean flux divergence at levels above the highest terrain. At lower levels, the mean flux divergence over that portion of the area where the pressure surface is actually above ground level will obviously be greater than the total areal average.

January and July profiles for the two areas are shown in figure 1. The January profile for the eastern area shows negative flux divergence at all levels. The maximum near 925 mb. coincides with the level of maximum inflow from the south. No strong increase in divergence is found at the level of maximum outflow on the east coast (750–800 mb.), as this outflow is more than offset by mean inflow through the three remaining boundaries.

The January profile for the western area differs from that in the east in some important respects. With the exception of the 1000-mb. level, the region below 850 mb. is found to be divergent. Examination of the flux along the boundary reveals outflow in the lower levels east of the Continental Divide and also into the Gulf of California which more than offsets the inflow across the Pacific Coast. Since both the Colorado River Basin and the area east of the Continental Divide are isolated from the remainder of the western region by mountains which rise above the 850-mb. level, the outflow source appears to be one or more of the following: evapotranspiration or decreased atmospheric storage within these isolated regions, a downward flux from the higher levels, or evaporation from falling precipitation. The mean flux divergence over the western region becomes negative at 850 mb. and maintains a rather constant negative value to 400 mb. The flux convergence above 650 mb. is significantly greater than that found over the eastern area.

The July profiles are similar in both east and west, but differ markedly from those found during January. Strong net inflow in the lower levels is capped by divergence at higher levels. In the east, the maximum inflow occurs between 950 and 1000 mb., and again coincides closely with the level of maximum inflow from the Gulf of Mexico. In addition, the level of maximum net outflow (between 800 and 900 mb.) now coincides with the level of maximum transport across the east coast.

Low level convergence is found throughout a deeper layer over the western area. This is probably a consequence of the extensive areas of high terrain, and the variable elevation of the ground. The computed magnitude of this convergence may be somewhat excessive between 850 mb. and 950 mb. because of a probable excess of low level inflow across 100°W. associated with systematic divergence errors east of the Rocky Mountains (see section 5). Similarly, the low level convergence of the eastern region may actually extend through a somewhat deeper layer than indicated by the computation. The high level flux divergence is located at considerably higher elevations over the west, in a manner similar to the high level convergence pattern in winter.

The contribution to the total integrated flux divergence from each 50-mb. layer is given in table 1. Values above 400 mb. were obtained by assuming a linear decrease of

TABLE 1.—*Vapor flux divergence. Units: gm.(cm.² mo.)*$^{-1}$

Pressure (mb.)	West		East		Total	
	Jan.	July	Jan.	July	Jan.	July
1000–950	+0.02	−0.87	−0.43	−1.02	−0.21	−0.95
950–900	+ .14	−1.45	− .69	+.03	− .28	− .70
900–850	+ .18	−1.15	− .59	+.60	− .21	− .25
850–800	+ .03	− .32	− .49	+.66	− .23	+ .18
800–750	+ .16	+ .10	− .44	+.51	− .30	+ .31
750–700	− .26	+ .37	− .35	+.33	− .31	+ .35
700–650	− .29	+ .47	− .29	+.28	− .29	+ .37
650–600	− .33	+ .53	− .22	+.24	− .28	+ .38
600–550	− .31	+ .59	− .14	+.12	− .22	+ .35
550–500	− .27	+ .50	− .13	+.11	− .20	+ .30
500–450	− .25	+ .35	− .12	+.10	− .18	+ .22
450–400	− .21	+ .21	− .08	+.09	− .15	+ .15
(400–250)	(− .26)	(+ .25)	(− .08)	(+.12)	(− .17)	(+ .18)
$\langle \nabla \cdot \overline{\mathbf{Q}} \rangle$	−1.97	− .42	−4.05	+2.17	−3.03	+ .89
$\langle \Delta W \rangle$ gm. cm.$^{-2}$	+ .5	+ .3	+ .2	+.3	+ .3	+ .3
Contribution above 500 mb.	−.72	+ .81	− .28	+ .31	−.50	+ .55

divergence to zero at 250 mb. Contributions from the layer below 1000 mb. are small, and are not included.

The July profiles may be compared with the June–August 1954 results of Hutchings [10] for southern England. He found strong convergence below 850 mb., and divergence at all higher levels up to 350 mb. His values were much larger than the July values over North America, with peak values of 10^{-1} gm.(cm.² mb. mo.)$^{-1}$ for the low level convergence and 3.3×10^{-2} gm.(cm.² mb. mo.)$^{-1}$ for high level divergence.

The contribution to the total vertically integrated divergence from the layers above 500 mb. is surprisingly large over the higher terrain of the western region, and it is quite apparent that significant systematic errors will arise in the computed mean monthly divergence if these layers are not included in the vertical integration. On the other hand, such errors would apparently reverse sign with the season, a consequence of the fact that the higher layers are convergent in winter and divergent in summer. Consequently such errors will have the effect of damping the actual seasonal variation of flux divergence. Since these seasonal errors will tend to cancel, the average annual error may not be large. The contribution from the layers above 500 mb. follows a similar pattern in the east, but here amounts to only 7 to 15 percent of the total integrated flux divergence.

Given these data, together with an estimate of the rate of evapotranspiration from the surface of the earth, one can estimate the vertical vapor flux through the lower atmospheric layers in those cases where condensation is not a significant factor. Computations of the vertical flux were made for the eastern area at a few of the lower levels, assuming no condensation losses and no changes in atmospheric storage in the layers. Estimates of evaporation from the earth's surface were based on the water balance computations to be discussed later in this paper, and are listed in table 2 as the flux from the surface.

It is interesting to note that even with the strong low level convergence observed in July, the vertical vapor flux at 900 mb. differs little from the surface evaporation

TABLE 2.—*Eastern area—computed vertical water vapor flux (assuming no condensation or atmospheric storage changes). Units: gm./cm.² mo.*

	January	July
Surface	2.5	9
950 mb	3.0	10
925 mb	3.5	10
900 mb	------	10

rate. Hutchings [10] computed a vertical transport of around 12½ gm.(cm.² mo.)$^{-1}$ through the 950-mb. level. Of this amount he estimated 6 gm.(cm.² mo.)$^{-1}$ was transported by the large-scale vertical motions, and the remainder by convection and small-scale turbulence. Evaporation was estimated at around 8 gm./mo. Rainfall during the 3-mo. period that he investigated was abnormally high (140 percent of normal). It seems probable that large-scale vertical motion plays a more important role in the summertime vertical vapor flux over England (particularly during the excessively wet summer of 1954) than it does over the United States south of 47.5°N.

The values of the low level vertical flux shown for January are around one-third of those found in July. Because of the neglect of condensation these values may be slight overestimates of the actual vertical flux.

5. CHARACTERISTICS OF THE FLUX DIVERGENCE MAPS

The regional water balance computations to be discussed in the next section yield information concerning the accuracy of the computed mean flux divergence averaged over relatively large areas. More detailed information concerning the characteristics of the systematic error field can be obtained if the flux divergence is computed with a higher degree of resolution. Such computations were made, as previously noted, using a 2.5°×2.5° grid south of 57.5°N. and a 2.5° lat.×5.0° long. grid north of 57.5°N. Certain of these analyses are shown in figures 2–6.

A critical examination of figures 2–4 in light of what is known of the evaporation and precipitation patterns over the area leaves little doubt that an error pattern of considerable magnitude does indeed exist. These errors appear to be particularly pronounced on the summer map.

The extent to which the errors obscure the true pattern is probably best illustrated by the mean annual divergence map (fig. 2). This map apparently captures the broad-scale features of the divergence pattern. The Central American Sea is shown as being primarily divergent. Convergence is the rule over the continent, with the expected large values on the north Pacific Coast and in the southeastern United States. However, the gradients in many areas and the magnitude of many of the major features on this map cannot be supported by independent hydrological data, and in many cases are undoubtedly in error.

Problems in the divergence distribution over the Central American Sea were anticipated, even though

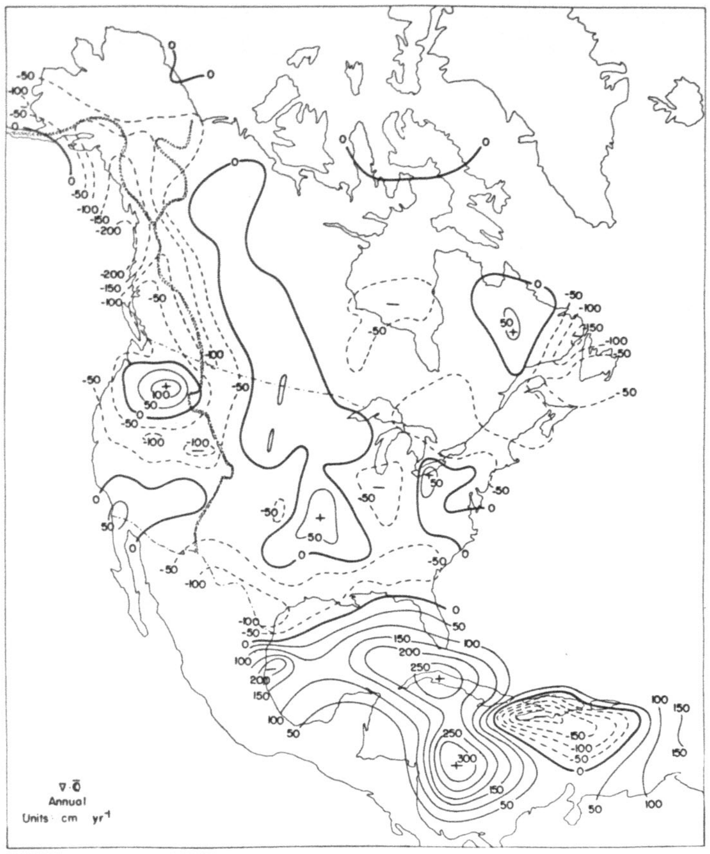

FIGURE 2.—Computed mean annual divergence of the vertically integrated total water vapor flux. May 1961–April 1963. Units: cm. yr.$^{-1}$ Values are an average of 00 GMT and 12 GMT observations.

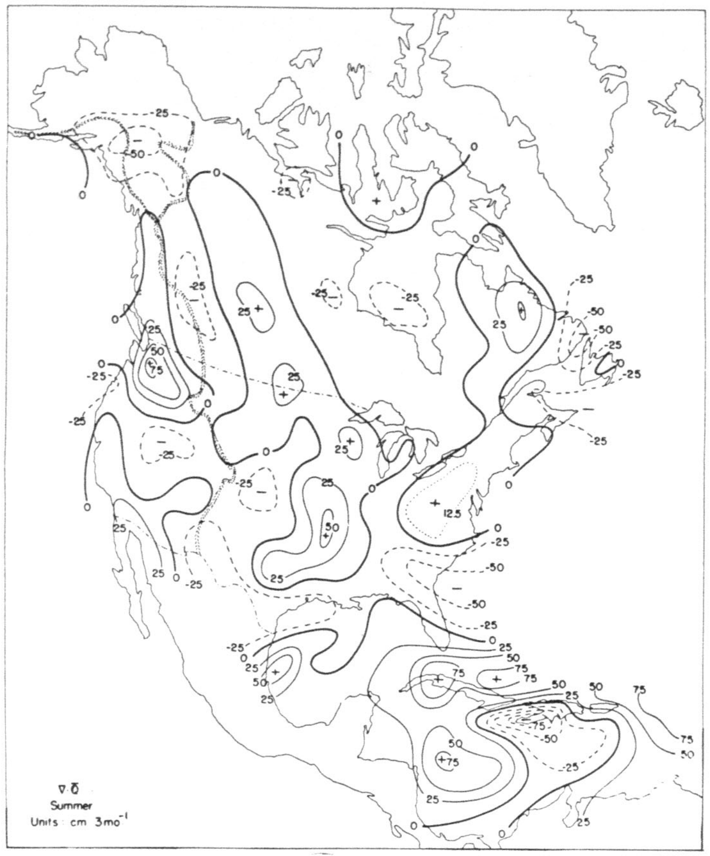

FIGURE 3.—Computed mean seasonal divergence of the vertically integrated total water vapor flux. Summer (June–August). Units: cm. (3 mo.)$^{-1}$. Values are an average of 00 GMT and 12 GMT observations.

efforts were made to produce a smooth field. The flux data from Kingston, Jamaica, appeared to be strongly influenced by local conditions, particularly during winter, and the very strong gradient between convergence in the northeastern Caribbean and divergence to the west may be due, in part, to improper interpretation of these data.

Data from the missile range stations in the Bahamas were not available with sufficient regularity to be of use during this 2-yr. period, nor were any data available from Havana. Consequently, the distribution of divergence over Cuba and the Florida Straits, and in the area to the east of the Greater Antilles is unreliable. Furthermore, data over Florida, and computations on a 2.5° grid, are not sufficient to adequately resolve differences between values of divergence over the peninsula and over the surrounding waters of the Gulf of Mexico and the Atlantic.

Some of the features along the edges of the continent are due to uncertainties of analysis, but most of the large-scale pattern over North America is well established by the data. Questionable features over the continent include:

1) The intense area of divergence over the northwestern United States and the excessive convergence to the south of this area.

2) The elongated area of divergence parallel to and just to the east of the Continental Divide, extending from the Yukon Territory almost to the Gulf Coast.

3) The strong convergent area over southern Texas.

4) The strong convergence over and just to the east of the Continental Divide.

5) The area of convergence extending from south of Lake Michigan, northward, then eastward through Ontario. It is the intensity of the convergence in this area which is in question.

6) The divergent area extending from Lake Erie to Hatteras.

7) The divergent area over northeastern Quebec and northern Labrador.

8) The convergence maximum over Hudson Bay.

9) The maxima over Labrador and Newfoundland. Again in (8) and (9), it is the magnitude that is primarily in question.

Examination of the seasonal analyses (including those for spring and autumn not shown here) and annual mean maps for the 2 individual yr. (not shown here) revealed the following facts:

1) All of the previously described features appeared on individual annual mean maps in approximately the same geographical locations but varied in intensity.

2) The strong convergence over Hudson Bay did not appear in winter and spring and the divergence over northeastern Quebec and northern Labrador did not appear in winter. All other features were recognizable on each seasonal map.

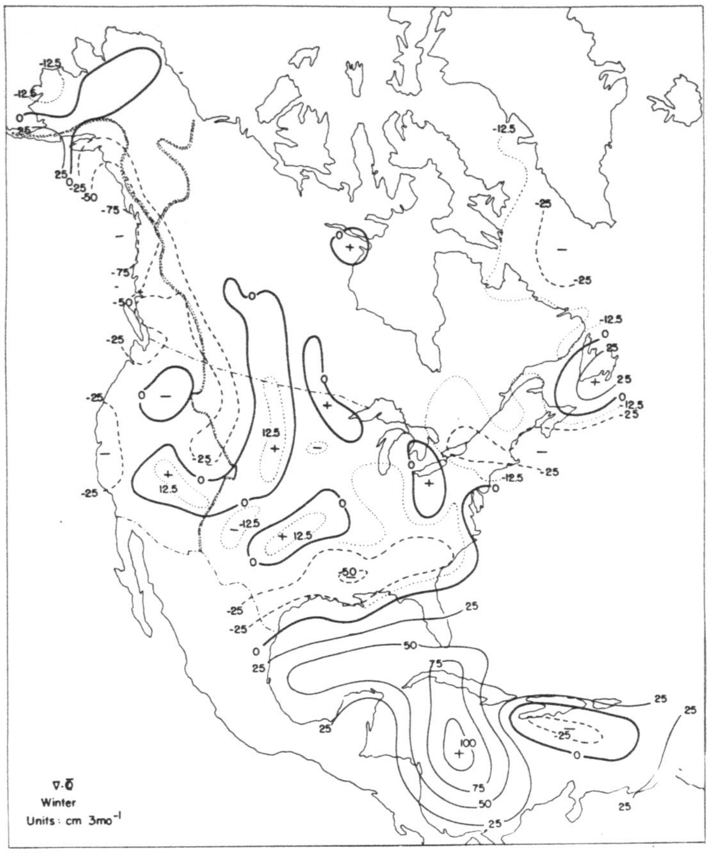

FIGURE 4.—Same as figure 3, but for winter (December–February).

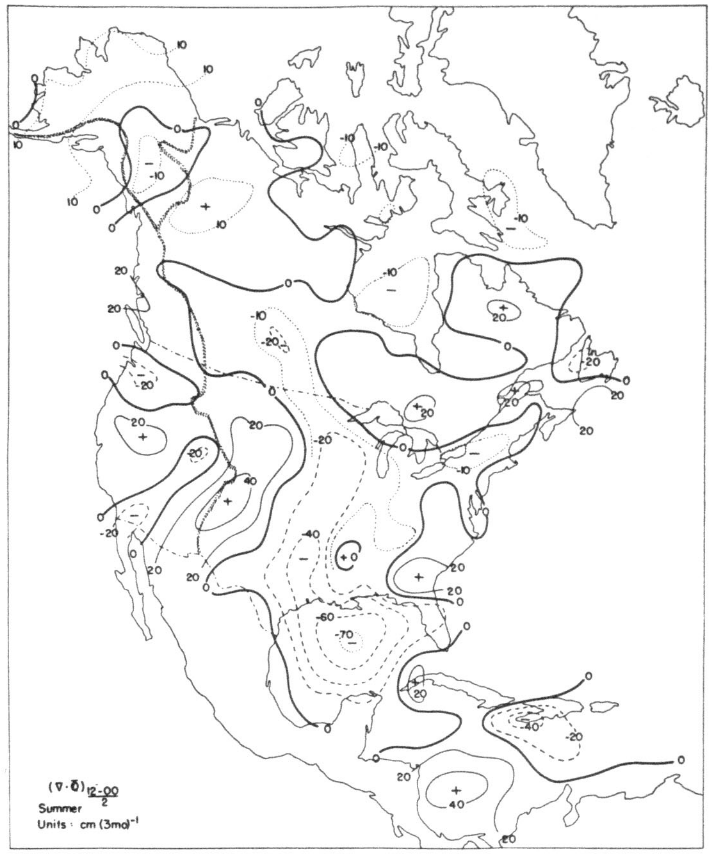

FIGURE 5.—Mean difference (12 GMT–00 GMT)/2, of the divergence of the vertically integrated total water vapor flux. Summer (June–August). Units: cm. (3 mo.)$^{-1}$.

It therefore appears reasonable to conclude that the error pattern observed during this 2-yr. period was primarily systematic in nature, appearing each year and, for the most part, in all seasons (except north of 50°). In this regard it should be noted that several of the major features appearing on these maps are also apparent in the less detailed analysis of 1958 data by Starr, Peixoto, and Crisi [23]: notably the convergent area over south Texas and the southern Rockies, the belt of divergence east of the Continental Divide, and the excessive convergence over the Canadian Rockies.

Further investigation is required before one can determine with some degree of certainty the primary reasons for these errors. However, because of diurnal variations in the vapor flux, which were discussed in Part I of this paper, the use of only twice-daily observations to define the mean daily flux may be one source of error. In this regard, figures 5 and 6 clearly show that the mean flux divergence, as well as the flux itself, exhibits a large diurnal variability. Comparison of figures 3 and 5 and figures 4 and 6 shows that during summer, and during winter south of 40°N., the diurnal variations in the flux divergence are of the same order of magnitude as the computed mean flux divergence itself.

The summertime pattern of diurnal change is dominated by the effects of the large-scale oscillation over eastern North America and the Gulf of Mexico. Because of the rapid decrease of specific humidity with height, the diurnal variation of $\nabla \cdot \mathbf{Q}$ is normally of the same sign as the velocity divergence in the lower troposphere. Thus the decrease in vapor flux convergence over the Rockies and high plains and the increase in convergence over the Mississippi Valley from 00 to 12 GMT are broadly consistent with the low level convergence patterns found by Bleeker and Andre [4] and the vertical motion field found by Curtis and Panofsky [7]. The greatest changes in divergence are computed over the Gulf of Mexico, where differences between 00 and 12 GMT reach values in excess of 50 gm.(cm.^{2} mo.)$^{-1}$. In the course of his investigation of the diurnal flux variations over the Gulf of Mexico, Hastenrath [9] computed the difference between the wind divergence at 00 and 12 GMT for the months of January, April, July, and October 1960. With the exception of April, these computations also show, in agreement with our findings, significantly stronger low level convergence over the Gulf of Mexico at 12 GMT.

Wintertime differences between the 12 GMT and 00 GMT flux divergence are much reduced but many of the features of the summer pattern can still be recognized. The changes over the Gulf of Mexico are still quite pronounced but the pattern over the Plains, although still identifiable, is quite weak. The pattern of variations north of 52.5°N. has almost completely disappeared, except in the area over Alaska and the Yukon, and there the summertime pattern is reversed.

Figures 2–4 give some clue as to the results one might expect when computing the water balance for large drainage areas. The distance between divergence centers of like sign varies considerably, but averages around 1500 km. over the United States and southern Canada.

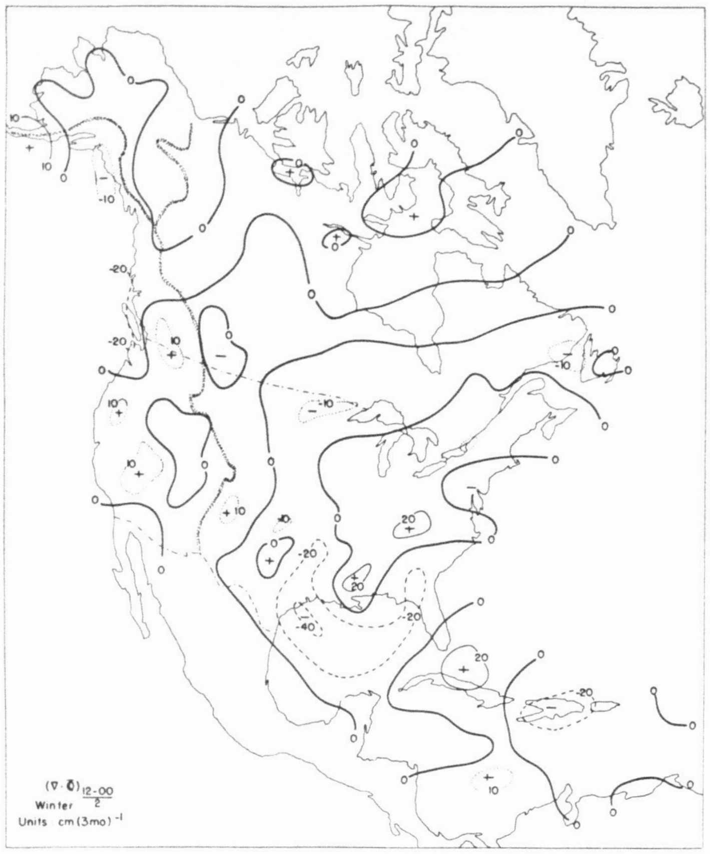

FIGURE 6.—Same as figure 5, but for winter (December–February).

If, for the moment, we assume that these major features are primarily noise, then one would expect a considerable reduction in the systematic error when dealing with averages over roughly circular or square areas of the order of 20×10^5 km.^{2} Water balance computations for 16 areas varying in size from 86.5×10^5 km.^{2} down to 2.5×10^5 km.^{2}, which were made prior to the construction of these maps, appear to support this conclusion (Rasmusson [20]). Good results were obtained for areas of about 20×10^5 km.^{2} and larger. Results for areas of 10–20×10^5 km.^{2} were fairly good but as the area was decreased to less than 10×10^5 km.^{2}, the results became much more erratic.

6. LARGE-SCALE WATER BALANCE COMPUTATIONS

UNITED STATES AND SOUTHERN CANADA

Rasmusson [20] has previously discussed some general aspects of the continental water balance and the water balance of northern North America. However, computation of surface and subsurface storage changes over the continent as a whole was precluded by inadequate streamflow data. On the other hand, data are sufficient for such computations over the United States and portions of southern Canada. This area consists of the combined Western, Central Plains, and Eastern Regions of figure 7, a total area of 86.5×10^5 km.^{2} from which all streamflow is measured.

Mean monthly values of the computed difference between precipitation and evapotranspiration $\langle\overline{P-E}\rangle$, runoff $\langle\overline{R_o}\rangle$, and storage change $\langle\Delta S\rangle$ for the total area are shown in figure 8. Annual values are given in table 3. These values represent averages for an area over which mean annual precipitation varies locally from less than 15 cm. to over 250 cm., and mean annual runoff varies from 0 to over 100 cm. (Miller, Geraghty, and Collins [16]). The seasons of highest and lowest flow differ locally, but for the area as a whole the maximum outflow occurred in spring and the minimum in the fall. Mean monthly runoff ranged from 0.8 to 2.6 cm. during the 2 yr. investigated.

The pattern of wintertime streamflow was considerably different in each of the 2 yr. Marked increases from the fall minimum were observed during the first year, while little or no recovery took place during the second year. The relatively low flow during the second winter was primarily the result of unusually cold and dry conditions over the eastern part of the continent.

$\langle\overline{P-E}\rangle$ shows a more irregular pattern and greater seasonal changes than does the streamflow. Maximum values occur during the winter, minimum values during the summer. In contrast to the northern sections of the continent, where precipitation exceeded computed evapotranspiration throughout the year, one finds an excess of evapotranspiration during the 3 summer mo. Computations over the portion of the area east of the Continental Divide, using 5 yr. of data, indicated a smaller excess of summertime evaporation over precipitation than was computed from the 2-yr. data sample.

Since the difference between $\langle\overline{P-E}\rangle$ and $\langle\overline{R_o}\rangle$ changes sign from winter to summer, there must be an accumulation of water over the continent during the late fall, winter, and early spring, which is lost again during the warmer months of the year. The computed seasonal change in storage is also shown in figure 8. These values represent the total change in storage from May 1, 1961.

The characteristics of the computed seasonal storage change agree qualitatively with what is known of this quantity. Soil moisture, as well as the water table, reach their highest values over most of the area in spring, and surface storage in the form of snow reaches a maximum in late winter and early spring. Late spring and summer mark a period of high evapotranspiration and decrease in storage. The lowest values of soil moisture, water table, and streamflow occur in late summer or early fall over most of the area.

Van Hylckama [28] has estimated the storage over the continents using the empirical techniques of Thornthwaite (Mather [15]). Mean monthly values were computed for the land area within each 10°×10° region of the earth. The average monthly storage changes as computed for a combination of areas which approximates the United States and southern Canada (30°–50°N., 70°–130°W. plus 50°–60°N., 100°–100°W.) are shown in figure 9, along with the results of this investigation. Van Hylckama's estimates were taken to represent storage on the 15th of each month.

The two curves are nearly in phase, although the maximum and minimum values computed by the water vapor balance equation appear to lag those of Van Hylckama by about ½ mo. On the other hand, the amplitude of the storage curves differs by more than a factor

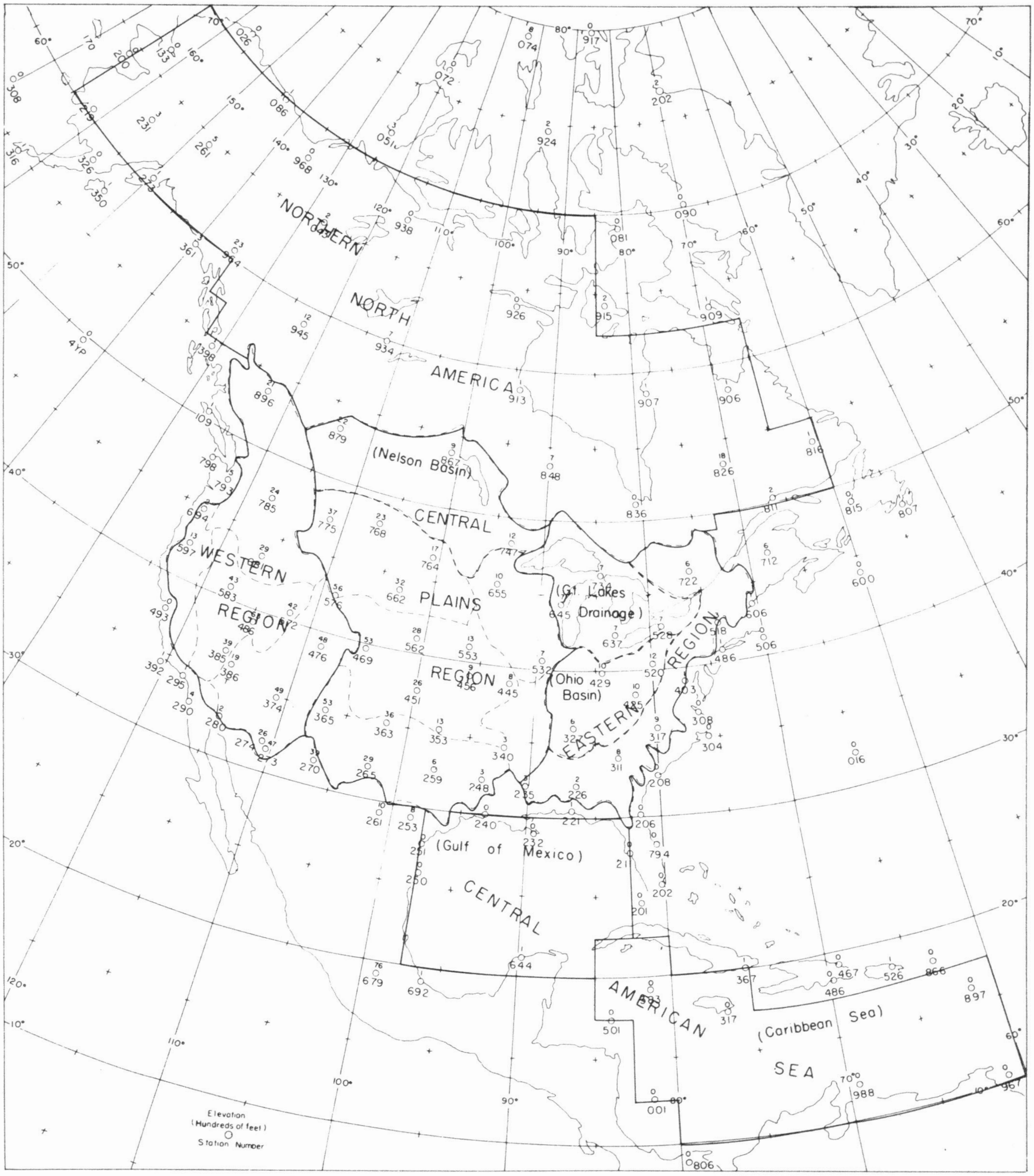

FIGURE 7.—Regions of water balance computations.

of two. A systematic underestimation of the moisture flux and consequent underestimation of flux divergence might be suggested as a possible reason for this difference, but since the mean annual flux divergence over the area is negative, this would lead to a sizable systematic overestimation of the storage loss. Such does not appear to be the case, since only a small net storage change was computed during the 2-yr. period. A flux error which varies systematically throughout the year could also produce erroneous values of seasonal storage change, and still give correct year-to-year changes. As was shown in section 4, this type of error can arise if the vertically integrated flux does not include the contribution from the layers above 500 mb. This, however, does not appear to be a factor in the present investigation since all available data up to 300 mb. were used.

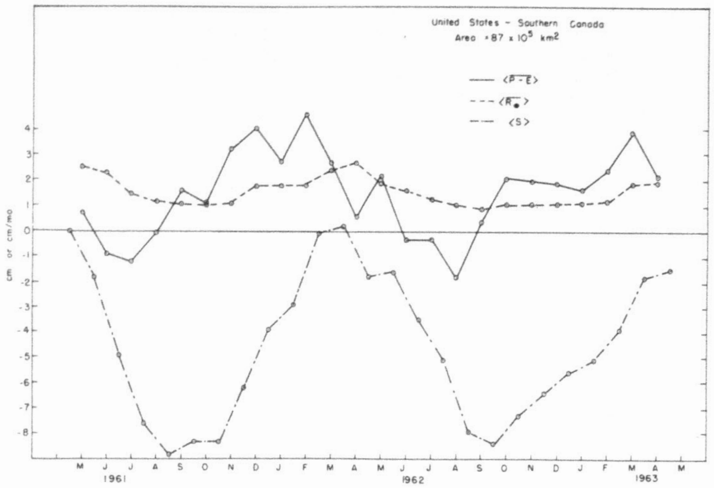

FIGURE 8.—Water balance of the United States-southern Canada. Surface and subsurface storage, $\langle S \rangle$, represents the change from May 1, 1961. The difference between precipitation and evaporation, $\langle \overline{P-E} \rangle$, is computed from the water vapor balance equation.

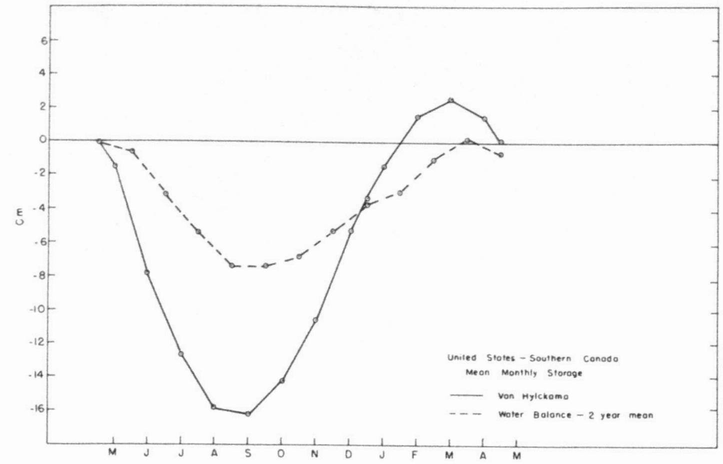

FIGURE 9.—Computed mean monthly surface and subsurface storage changes.

TABLE 3.—*United States and southern Canada. Area=86.5×10⁵ km.² Units: cm. yr.⁻¹*

	May 1961–Apr. 1962	May 1962–Apr. 1963
$\langle \overline{P-E} \rangle$	18.8	16.0
$\langle \overline{R_o} \rangle$	20.6	15.7
$\langle \Delta S \rangle$	−1.8	+0.3

Since the amplitude of the annual storage curve was not very different during each of the 2 yr. studied, it seems probable that the results of the water balance computation are a reasonable estimate of the long-term mean seasonal storage changes. It is suggested that the Thornthwaite method tends to underestimate $\overline{P-E}$ during summer and overestimates it during winter, and that this accounts, at least in part, for the greater amplitude of Van Hylckama's storage curve. Evidence for this statement will be presented later in this paper.

Since there is a difference in the land area of the Northern and Southern Hemisphere, these seasonal changes in storage represent a substantial seasonal shift of water from the oceans to the continents. Consequently, the total water content of the oceans is lowest in March and highest in October (Donn, Patullo, and Shaw [8]). The difference represents only a small contribution to changes in mean sea level and is thus difficult to estimate. Van Hylckama cites a calculation by Munk, using tidal gage data, which indicates an oceanic storage change from March to October of 0.50×10^{19} gm.(1.4 cm.). Van Hylckama himself computes a change of 0.75×10^{19} gm.(2.10 cm.). However, the comparison of his computed storage changes over the United States and southern Canada with the results from the vapor balance equation suggests that his value may be too high.

CENTRAL AND EASTERN NORTH AMERICA

Mean monthly water balance.—The combined Central Plains and Eastern Region (see fig. 7) was chosen for the most detailed study. Mean monthly precipitation for the entire area was estimated from zonal averages recorded in state climatological summaries, and from the precipitation maps of the Canadian Department of Transport Monthly Report. The period of investigation was extended back to May 1, 1958, by using the flux divergence computations of Bock, Frazier, and Welsh [5].

The lengthening of the period of investigation allows one to assume, as a first approximation, that the mean annual surface and subsurface storage change, averaged over the 5-yr. period, is zero. Any computed net storage change is then attributed to a systematic error in the evaluation of $\nabla \cdot \overline{\mathbf{Q}}$. A net increase in storage of 21 cm. was computed for the 5-yr. period. Since there is as yet no firm information concerning the seasonal distribution of systematic errors, a uniform correction of +0.35 cm./mo. was applied to the computed divergence in order to reduce the 5-yr. computed storage change to zero. This correction is relatively small, and the final results would not be strongly affected if, for instance, one were to apply the entire correction during the 6 summer or 6 winter mo.

Mean monthly values of runoff, precipitation, evapotranspiration, and storage are given in table 4 and in figure 10. Departures from the 5-yr. annual average values are given in table 5. The values of runoff are largely determined by the Eastern Region, which accounts for only 35 percent of the total area, but contributed 69 percent of the runoff. Total precipitation volume is slightly higher over the Central Plains Region, although average precipitation is significantly higher over the Eastern Region.

The storage curve for the area has much the same characteristics as that shown in figure 9, as would be expected. The computed minimum storage at the end of August and the maximum at the end of March differ by about 6 cm. Evapotranspiration during the winter ranges around 1½ to 2 cm./mo., while a summertime maximum of around 9 cm./mo. is computed in July. The seasonal march of evapotranspiration is very similar to that obtained for the entire continent by Benton and Estoque [3]. Precipitation exceeds computed evapotranspiration during all months except July and August, when the difference is hardly significant.

TABLE 4.—*Central Plains and Eastern Regions. Area=64×10^5 km.*[2] *Units: cm./mo. or cm. Computed mean monthly water balance components. (May 1958–Apr. 1963)*

	$\langle\overline{R_0}\rangle$	$\langle\overline{P}\rangle$	$\langle\overline{E}\rangle$	$\langle\Delta S\rangle$	$\langle S\rangle$*
Sept.	1.02	7.33	4.99	+1.32	1.32
Oct.	1.13	5.64	4.05	+ .46	1.78
Nov.	1.10	4.71	2.93	+ .68	2.46
Dec.	1.38	4.46	2.22	+ .86	3.32
Jan.	1.55	4.04	2.34	+ .15	3.47
Feb.	1.73	4.71	1.23	+1.75	5.22
Mar.	2.51	5.47	2.48	+ .48	5.70
Apr.	2.62	5.61	4.04	−1.05	4.65
May	2.29	7.93	6.27	− .63	4.02
June	1.61	8.69	8.05	− .97	3.05
July	1.44	9.09	9.34	−1.69	1.36
Aug.	1.22	7.16	7.30	−1.36	.00
Annual	19.60	74.74	55.24		

*As of end of month. Change from September 1.

We have chosen to compare our computed evapotranspiration and storage change with values obtained using two well-known and widely used estimates of evapotranspiration: those of Budyko [6] and C. W. Thornthwaite and Associates [25, 26]. Budyko's values must be obtained by interpolation from relatively small maps, but interpolation errors are probably less than ½ cm./mo. The Thornthwaite evapotranspiration and storage estimates were obtained from analyses over the area of interest, using their data for 497 stations distributed uniformly over the area. Storage changes were also computed using observed values of runoff and Budyko's values of evapotranspiration.

Figure 11 shows a comparison of the three storage estimates. The storage changes derived from Budyko's evapotranspiration estimates do not balance for the year, indicating that his value of mean annual evapotranspiration is around 3 cm. less than that required to obtain a balance with the measured precipitation during this 5-yr. period. Nevertheless, the values are in fair agreement with those obtained from the vapor balance equation, particularly if one allows for the 3-cm. imbalance. On the other hand, the Thornthwaite estimate follows a pattern which, as might be expected, is similar to the previously discussed estimate of Van Hylckama for the United States and southern Canada. His computed seasonal storage change is again more than double that obtained from the vapor balance equation. Significantly higher values of seasonal storage change were also obtained from the Thornthwaite data for each of the major subdivisions of the area (Central Plains Region, Eastern Region, Ohio Basin, and Great Lakes Drainage). Thus, the difference in the results obtained from the two methods appears to be quite systematic.

The basic reason for this difference is evident from figure 12. The Thornthwaite wintertime values of evapotranspiration are significantly lower and summertime values significantly higher than those obtained from the vapor balance equation. This pattern is, again, also found over the smaller subdivisions of the area. Since measured precipitation is used in both balance schemes, computations using the Thornthwaite data show a greater

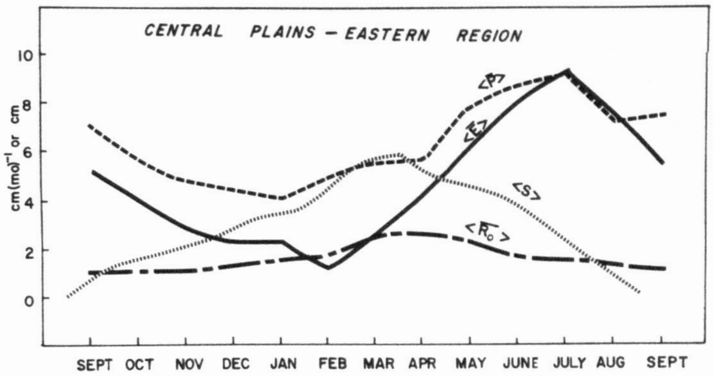

FIGURE 10.—Water balance: Central Plains and Eastern United States. May 1958–April 1963. Units: cm./mo.

TABLE 5.—*Central Plains and Eastern Regions. Computed departures from 5-yr. annual averages. Period: May 1958–April 1963. Units: cm./yr.*

Parameter	Year				
	1	2	3	4	5
$\langle\overline{P}\rangle$	−2.3	+5.3	−0.6	+1.9	−4.3
$\langle\overline{R_0}\rangle$	+0.0	+0.6	+0.0	+3.3	−3.9
$\langle\overline{E}\rangle$	+4.3	−2.8	+0.3	−0.4	−1.4
$\langle S\rangle$	−6.6	+7.5	−0.9	−1.0	+1.0
$\langle\overline{P-E}\rangle$	−6.6	+9.1	−0.9	+2.3	−2.9

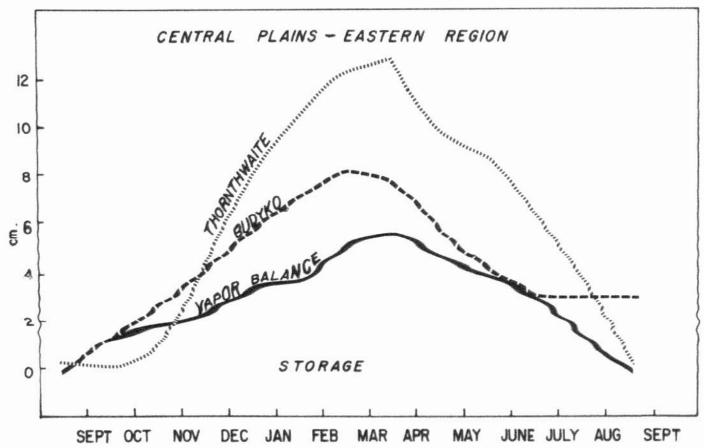

FIGURE 11.—Comparison of estimates of mean monthly surface and subsurface storage change: Central Plains and Eastern United States. May 1958–April 1963. Units: cm.

accumulation of storage during winter, and a greater loss during summer.

The Budyko values are more in line with those obtained from the vapor balance equation, although they show an interesting and rather consistent lag of around ½ to 1 mo.

Mean monthly variability.—The balance computations previously described provide a time series of 60 mean monthly values for each of the four hydrologic parameters. The annual march was removed from these data by subtracting the mean monthly values, and a few of the statistical properties of the resulting series were examined. Some of the results are given in tables 6 and 7.

The standard Chi-square goodness of fit test supported the null hypothesis of normality for each of the series.

Serial correlation coefficients, computed for each of the parameters, gave a lag 1 correlation which differed from zero by a statistically significant amount only in the case of runoff. The coefficients for these data, for lags 1 through 4, were 0.47, 0.44, 0.24, and 0.08, suggesting considerable persistence in the series. Because of this lack of independence of mean monthly values, no confidence limits were computed for relationships involving runoff.

Cross correlation coefficients between $\langle \overline{P} \rangle$ and $\langle \overline{R_o} \rangle$, and between computed $\langle \Delta S \rangle$ and $\langle \overline{E} \rangle$ did not differ from zero by a statistically significant amount for either lag zero or lag 1.

A significant relationship existed between measured $\langle \overline{P} \rangle$ and $-\langle \nabla \cdot \overline{\mathbf{Q}} + \Delta W \rangle$ (which represents the computed value of $\langle \overline{P-E} \rangle$) at lag zero. Sample correlation coefficients for the summer 6 mo. (May–October), winter 6 mo. and for the year were 0.83, 0.93, and 0.87 respectively. Thus, the wintertime variance of computed $\langle \overline{P-E} \rangle$ is almost entirely accounted for by variations in measured $\langle \overline{P} \rangle$ and even during the summer about 70 percent of the variance of computed $\langle \overline{P-E} \rangle$ could be accounted for by the variation of $\langle \overline{P} \rangle$. When one considers the fact that some error exists in both the evaluation of $\langle \overline{P} \rangle$ and $\langle \overline{P-E} \rangle$, and the improbability of any significant correlation between these errors, it seems clear that departures from normal of $\langle \overline{E} \rangle$ normally make only a small contribution to the departure of $\langle \overline{P-E} \rangle$ over this area.

The slope of the regression line (with $\langle \overline{P} \rangle$ as the abscissa) was 1.40 in summer and 1.05 in winter. A slope greater than 1.00 indicates a tendency for $\langle \overline{E} \rangle$ to decrease as $\langle \overline{P} \rangle$ increases, but during neither season does the slope differ from one by an amount statistically significant at the 5-percent level.

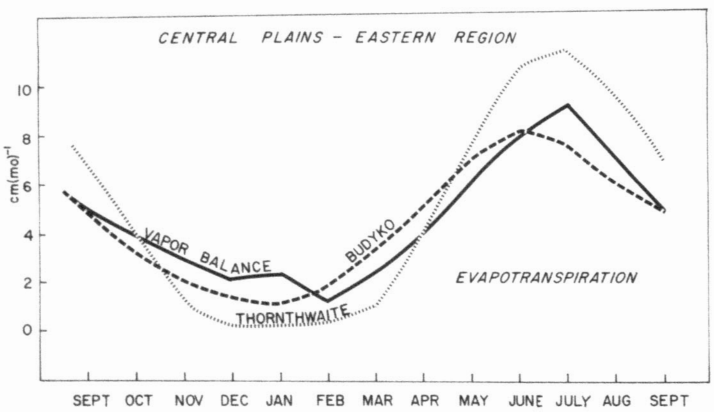

FIGURE 12.—Comparison of estimates of mean monthly evapotranspiration: Central Plains and Eastern United States. May 1958–April 1963. Units: cm./mo.

Computed values of $\langle \overline{E} \rangle$ and $\langle \Delta S \rangle$ were negatively correlated at lag zero, the coefficients being -0.85 in summer and -0.54 in winter. However, one cannot necessarily interpret this as an indication of a real relationship between these quantities. Errors in the measurement of $\langle \nabla \cdot \overline{\mathbf{Q}} \rangle$ will be reflected as errors of opposite sign when computing $\langle \overline{E} \rangle$ and $\langle \Delta S \rangle$, so at least part of the negative correlation arises simply from these errors.

Next to the correlation between $\langle \nabla \cdot \overline{\mathbf{Q}} + \Delta W \rangle$ and $\langle \overline{P} \rangle$, the correlation between $\langle \Delta S \rangle$ and $\langle \overline{P} \rangle$ at lag zero was the most significant of the relationships between the four parameters. Summer, winter, and annual coefficients of 0.78, 0.87, and 0.79 were obtained. Thus variations in $\langle \overline{P} \rangle$ accounted for about half the summertime variance in computed $\langle \Delta S \rangle$ and for more than three-fourths of its winter variance. The unexplained variance arises from variations in runoff not correlated with storage changes, and from errors in the evaluation of $\langle \nabla \cdot \overline{\mathbf{Q}} + \Delta W \rangle$. The slope of the regression line (1.41 during summer, 0.86 during the winter) does not differ from one by an amount statistically significant at the 5-percent level.

The picture of average conditions over the Central Plains and Eastern Regions which emerges from this analysis can be summarized as follows. Variations from mean monthly $\langle \overline{R_o} \rangle$ and $\langle \overline{E} \rangle$ are at best only weakly related to variations in $\langle \overline{P} \rangle$. Furthermore, the magnitude of runoff variations is relatively small. On the other hand, there is a much stronger relationship between variations in $\langle \overline{P} \rangle$ and $\langle \Delta S \rangle$. Because of this, variations in $\langle \overline{P} \rangle$ are reflected most strongly as compensating changes in storage. Thus, when dealing with mean conditions over the entire area, precipitation departure *alone* serves as a fairly good indicator of the quantitative departure from normal

TABLE 6.—*Sample standard deviation of the departure from the mean monthly value. May 1958–April 1963. Units: cm. mo.$^{-1}$*

	$\langle \overline{P-E} \rangle$	$\langle \overline{P} \rangle$	$\langle \overline{E} \rangle$	$\langle \Delta S \rangle$	$\langle \overline{R_o} \rangle$
Summer	1.55	0.92	0.93	1.66	0.35
Winter	1.41	1.25	.52	1.24	.36

TABLE 7.—*Departures from mean monthly values. Regression relationships for zero lag*

Y	$-\langle \nabla \cdot \overline{\mathbf{Q}} + \Delta W \rangle$[1]			$\langle \overline{E} \rangle$			$\langle \overline{R_o} \rangle$			$\langle \Delta S \rangle$		
X	$\langle \overline{P} \rangle$			$\langle \Delta S \rangle$			$\langle \overline{P} \rangle$			$\langle \overline{P} \rangle$		
	Summer	Winter	Annual	Summer	Winter	Annual	Summer	Winter	Annual	Summer	Winter	Annual
r	0.83	0.93	0.87	−0.85	−0.54	−0.75	0.00	0.65	0.37	0.78	0.87	0.79
95% confidence	.63	.84	.79	− .92	− .76	− .85				.55	.72	.67
limits for r	.91	.96	.92	− .69	− .16	− .60				.90	.93	.87
b	1.40	1.05	1.17	−0.48	−0.23	−0.39	0.00	0.19	0.12	1.41	0.86	1.06
95% confidence	.93	.83	.93	− .63	− .41	− .51				.84	.61	.77
limits for b	1.87	1.27	1.41	− .33	− .05	− .27				1.98	1.11	1.35

[1] Represents computed value of $\langle \overline{P-E} \rangle$.

storage change. The validity of this statement for the area studied does not necessarily imply its validity for smaller areas.

Some hint as to the magnitude of the random error in $\langle \nabla \cdot \overline{\mathbf{Q}} \rangle$ can be obtained from this analysis. If the correlation between $\langle \Delta S \rangle$ and $\langle \overline{E} \rangle$ was due entirely to errors in $\langle \nabla \cdot \overline{\mathbf{Q}} \rangle$, its standard error was around 0.8 cm./mo. during summer and 0.3 cm./mo. during winter. Alternately, one may assume that all the variance of $\langle \nabla \cdot \overline{\mathbf{Q}} + \Delta W \rangle$ which cannot be explained by variations in $\langle \overline{P} \rangle$ is due to errors in $\langle \nabla \cdot \mathbf{Q} \rangle$, rather than to real variability in $\langle \overline{E} \rangle$ and errors in estimating $\langle \overline{P} \rangle$ and $\langle \Delta W \rangle$. This gives standard errors for $\langle \nabla \cdot \overline{\mathbf{Q}} \rangle$ of 0.9 cm./mo. in summer and 0.5 cm./mo. in winter. Because of the assumptions involved, it is probable that these figures are overestimates of the actual error. All in all, it seems likely that the standard error in $\langle \nabla \cdot \overline{\mathbf{Q}} \rangle$ for this particular area lies around 0.50 cm./mo. in summer, and around 0.25 cm./mo. in winter.

7. SUMMARY AND CONCLUSIONS

Certain aspects of the water balance of the North American continent and two large subareas of the continent have been investigated, using atmospheric vapor flux data together with observed streamflow and precipitation.

The mean vertical distribution of the flux divergence was computed for the United States for the months of January and July. Strong flux convergence in the lowest 100 mb. and divergence through the remainder of the troposphere were found in July. Flux convergence was found throughout the troposphere over the eastern half of the area in January, with a maximum between 900 and 950 mb., while in the west convergence (with no particularly pronounced maximum) was found above 800 mb. with weak divergence below. Corresponding features of the profiles were found at higher elevations over the west, where the flux divergence above 500 mb. is quite significant.

Maps of the vertically integrated flux divergence for North America and the Central American Sea exhibit systematic diurnal variations and also a systematic error pattern, which are of relatively large scale and amplitude. The presence of the systematic errors is the primary factor that leads to a deterioration of the results of the balance computations as the size of the area over which averages are taken is decreased. Balance computations using 2 to 5 yr. of data gave good results for areas of about 20×10^5 km.^{2} or larger, but as the area was decreased to less than 10×10^5 km.^{2} the results became much more erratic.

In this paper we have reviewed primarily the results of very large-scale balance computations. A discussion of the results for the smaller areas will be published at a later date.

For the United States and southern Canada, computed values of $\langle \overline{P-E} \rangle$ coupled with observed streamflow from the area gave an average storage curve whose late summer minimum and spring maximum differed by around 7 cm. In contrast, Van Hylckama [28], using Thornthwaite's technique, computed a difference of around 19 cm. Further computations based on the vapor balance equation, for five areas of eastern North America, all yielded seasonal storage changes which were significantly less than those computed from the water balance data of Thornthwaite and Associates [25, 26]. This difference appears to arise primarily from a systematic overestimation of wintertime $\overline{P-E}$ and underestimation of summertime $\overline{P-E}$ when the Thornthwaite data are used.

We have briefly looked at some of the statistical properties of the departures from the monthly means. The results of this analysis, particularly the strong correlation between variations in precipitation and $\langle \nabla \cdot \overline{\mathbf{Q}} + \Delta \mathrm{W} \rangle$ indicate that month-to-month variations in $\langle \overline{P-E} \rangle$ are being computed accurately enough to yield useful information. Specifically, the results indicate that when dealing with averages over the entire area, the departure from normal precipitation alone serves as a fairly good quantitative indicator of the departure from normal storage change. This finding may have implications for long-range forecasting since it suggests that on the larger scale a good forecast of the departure from normal storage change is primarily dependent on a good forecast of the departure from normal precipitation, with evapotranspiration departures being of secondary importance.

The results of this study leave little doubt as to the advantages which can be gained when vapor flux data, along with standard surface hydrologic data, are applied to large-scale hydrologic investigations. Many results can be obtained for regions of good aerological data, such as North America, which are difficult, if not impossible, to obtain in any other way. It is important to point out, however, that the data must be used on a time and space scale that is compatible with the density, frequency, and quality of the available observations. There is little point in stubbornly attempting to apply divergence computations on a scale for which they are not suited. On the other hand, it should also be made clear that the results of this investigation by no means represent the ultimate in what can be obtained from the existing observational network. The formation of mean monthly values from daily analyses on a number of pressure surfaces together with the careful consideration of topography might well produce marked improvement in the results. Furthermore, a study of the individual terms which make up the mean monthly flux divergence at a particular level, i.e.

$$\nabla \cdot q\mathbf{V} = \nabla \cdot \overline{q'\mathbf{V}'} + \overline{\mathbf{V}} \cdot \nabla \overline{q} + \overline{q} \nabla \cdot \overline{\mathbf{V}}$$

might isolate the major sources of error in the computations. Results of this study suggest that errors in the evaluation of the divergence of the mean monthly wind may contribute, through the third term on the right, a large part of the systematic error in $\nabla \cdot \overline{\mathbf{Q}}$. One may be able to devise rational smoothing techniques for the mean monthly wind field that will significantly reduce this error.

ACKNOWLEDGMENTS

Part of the research described in this paper was performed while I was on assignment at M.I.T., and I am deeply indebted to Professor Victor P. Starr for his guidance and encouragement during that phase of the investigation. The data processing was accomplished through the efforts of Mr. Howard Frazier of Travelers Research Center, Mr. Salmon Seroussi of Mitre Corporation, Mrs. Judy Copeland and Miss Judy Roxborough of M.I.T. Calculations were preformed at the Geophysical Fluid Dynamics Laboratory, Mitre Corporation, and the M.I.T. Computation Center.

The help of Miss I. Kole and Mr. R. Stoner in drafting the figures and Mrs. C. Bunce in typing the manuscript is appreciated.

I wish to express my appreciation to ESSA, and particularly to Dr. Joseph Smagorinsky for the opportunity to perform this research. Thanks are also due to Dr. A. Oort who reviewed the original manuscript.

That portion of the research performed at M.I.T. was supported by the National Science Foundation under grant numbers GP-820 and GP-3657. Additional support was received from the U.S. Air Force 433L program and Mitre Corporation.

REFERENCES

1. W. C. Ackerman, "Committee on Status and Needs in Hydrology: A Look at Data and Instrumentation," *Transactions of the American Geophysical Union*, Vol. 46, No. 4, Dec. 1965, pp. 700–715.
2. D. Atlas, "Model Atmospheres for Precipitation," *Handbook of Geophysics and Space Environment*, U.S. Air Force Cambridge Research Laboratory, Bedford, Mass., 1965, Chapter 5, pp. 6–11.
3. G. S. Benton and M. A. Estoque, "Water Vapor Transfer Over the North American Continent," *Journal of Meteorology*, Vol. 11, No. 6, Dec. 1954, pp. 462–477.
4. W. Bleeker and M. J. Andre, "On the Diurnal Variation of Precipitation, Particularly Over the Central U.S.A.," *Quarterly Journal of the Royal Meteorological Society*, Vol. 77, No. 332, Apr. 1951, pp. 260–271.
5. P. Bock, H. M. Frazier, and J. G. Welsh, "Moisture Flux Over North America: Part II. Analysis of Monthly and Longer Term Means of the Flux Divergence of Moisture Over Northern America, May 1958–April 1963," *Final Report*, Contract CWb-11313, Travelers Research Center, Inc., Hartford, Conn., 1966, 97 pp.
6. M. I. Budyko (Editor), *Atlas Teplovogo Balansa Zemnogo Shara*, U.S.S.R. Glavnaîa Geofizicheskaîa Observatorîa, Moscow, 1963, 69 pp. (I. A. Donehoo (Translator of text), *Guide to the Atlas of the Heat Balance of the Earth*, WB/T-106, U.S. Weather Bureau, Washington, D.C., Dec. 1964, 25 pp.
7. R. C. Curtis and H. A. Panofsky, "The Relation Between Large Scale Vertical Motion and Weather in Summer," *Bulletin of the American Meteorological Society*, Vol. 39, No. 10, Oct. 1958, pp. 521–531.
8. W. L. Donn, J. G. Pattulo, and D. M. Shaw, "Sea Level Fluctuations and Long Waves," *Research in Geophysics*, Vol. II, M.I.T. Press, Cambridge, Mass., 1964, pp. 243–269.
9. S. L. Hastenrath, "Diurnal Fluctuations of the Atmospheric Moisture Flux in the Caribbean and Gulf of Mexico Area," *Journal of Geophysical Research*, Vol. 72, No. 16, Aug. 1967, pp. 4119–4130.
10. J. W. Hutchings, "Water Vapor Flux and Flux Divergence Over Southern England: Summer 1954," *Quarterly Journal of the Royal Meteorological Society*, Vol. 83, No. 1, Jan. 1957, pp. 30–48.
11. H. J. aufm Kampe and H. K. Weickmann, "Physics of Clouds," *Meteorological Monographs*, American Meteorological Society, Boston, Mass., Vol. 3, No. 12–20, July 1967, pp. 182–225.
12. M. A. Kohler and M. M. Richards, "Multicapacity Basin Accounting for Predicting Runoff From Storm Precipitation," *Journal of Geophysical Research*, Vol. 67, No. 13, Dec. 1962, pp. 5187–5197.
13. J. A. La Rue and R. J. Younkin, "Large-Scale Precipitation Volumes, Gradients, and Distribution," *Monthly Weather Review*, Vol. 91, No. 8, Aug. 1963, pp. 393–401.
14. D. H. Lufkin, "Atmospheric Water Vapor Divergence and the Water Balance at the Earth's Surface," *Final Report, General Circulation Project*, Massachusetts Institute of Technology, Cambridge, Mass., Dec. 1959, pp. 50–93.
15. J. R. Mather, "The Climatic Water Balance," *Technical Note* No. 1, Contract AF 49(638)-887, C. W. Thornthwaite and Associates, Laboratory of Climatology, Centerton, N.J., 1961, pp. 252–264.
16. D. W. Miller, J. J. Geraghty, and R. S. Collins, *Water Atlas of the United States*, (2d Edition), Water Information Center, Port Washington, N.Y., 1963, 40 plates.
17. E. Palmén, "Computation of the Evaporation Over the Baltic Sea From the Flux of Water Vapor in the Atmosphere," *International Association of Scientific Hydrology, Publication* No. 62, 1963, pp. 244–252.
18. J. P. Peixoto, "O Campo da Divergência do Transporte do Vapor de Água na Atmosfera" [Divergence Field of the Transport of Water Vapor in the Atmosphere], *Faculdade de Ciências, Revista*, Ser. 2A, B, Vol. 7, Universidade Lisbon, 1959/1960, pp. 25–56.
19. E. M. Rasmusson, "Atmospheric Water Vapor Transport and the Water Balance of North America: Part I. Characteristics of the Water Vapor Flux Field," *Monthly Weather Review*, Vol. 95, No. 7, July 1967, pp. 403–426.
20. E. M. Rasmusson, "Atmospheric Water Vapor Transport and the Hydrology of North America," *Planetary Circulations Project Report* No. A-1, Massachusetts Institute of Technology, Cambridge, Mass., May 1966, 170 pp.
21. G. I. Roden, "On River Discharge Into the Northeastern Pacific Ocean and the Bering Sea," *Journal of Geophysical Research*, Vol. 72, No. 22, Nov. 1967, pp. 5613–5629.
22. V. P. Starr and J. P. Peixoto, "On the Global Balance of Water Vapor and the Hydrology of Deserts," *Tellus*, Vol. 10, No. 2, May 1958, pp. 189–194.
23. V. P. Starr, J. P. Peixoto, and A. R. Crisi, "Hemispheric Water Balance for the IGY," *Tellus*, Vol. 17, No. 4, Nov. 1965, pp. 463–472.
24. L. R. Struzer, I. N. Nechaev, and E. G. Bogdanova, "Sistematicheskie Pogreshnosti Izmereniîa Atmosfernykh Osadkov" [Systematic Errors in Precipitation Measurements], *Meteorologiya i Gidrologiya* [Studies in Meteorology and Hydrology], No. 10, Clearing House for Federal and Technical Information, Washington, D.C., Dec. 1965, pp. 50–54.
25. C. W. Thornthwaite and Associates, "Average Climatic Water Balance Data of the Continents: Part VI. North America (Excluding United States)," *Publications in Climatology*, Laboratory of Climatology, Centerton, N.J., Vol. 17, No. 2, 1964, pp. 235–414.
26. C. W. Thornthwaite and Associates, "Average Climatic Water Balance Data of the Continents: Part VII. United States," *Publications in Climatology*, Laboratory of Climatology, Centerton, N.J., Vol. 17, No. 3, 1964, pp. 419–615.
27. C. W. Thornthwaite and F. K. Hare, "The Loss of Water to the Air," *Meteorological Monographs*, American Meteorological Society, Boston, Mass., Vol. 6, No. 28, July 1965, pp. 163–180.
28. T. E. A. Van Hylckama, "The Water Balance of the Earth," *Publications in Climatology*, Vol. 9, No. 2, Laboratory of Climatology, Drexel Institute of Technology, Centerton, N.J., 1956, pp. 57–105.
29. L. L. Weiss and W. T. Wilson, "Precipitation Gage Shields," *Transactions of the International Union of Geodesy and Geophysics, Association of Scientific Hydrology, Toronto, September 3–14, 1957*, Vol. 1, 1958, pp. 462–484.

[*Received January 31, 1968*]